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Temperature, humidity and cloud near fronts in the middle and upper troposphere

by

J. S. Sawyer

(Meteorological Office)

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Introduction

Although from the initial introduction of the idea of a front into meteorology (Bjerknes, 1918) the concept has been three-dimensional, nevertheless until recent years the weather forecaster has not been accustomed either to take into account the observed structure of fronts in the middle or upper troposphere, or even to relate the distribution of cloud at these levels to the position of the front in the upper troposphere. The brilliant analysis techniques of indirect aerology introduced by T. Bergeron (1928) are three-dimensional in concept, but still only relate phenomena of the surface weather map, and it is this chart which has been, and still is, the forecaster's main tool in deciding the distribution of cloud and rain.

On the other hand the present network of upper air soundings enables reliable charts to be drawn for any selected level in the troposphere over much of the northern hemisphere, and attention has recently been directed to the well-defined nature of the frontal zone which can be identified upon these charts and, in particular, at the 500 mb level (Palmén, 1951). Godson, and others (Godson, 1950) have gone further and suggested that an analysis of the three-dimensional shape of frontal surfaces should be made regularly by means of frontal contour charts and should form part of the routine at major forecast offices.

It therefore becomes desirable to study the character of the frontal zone in mid-troposphere, and the relation of the mid-troposphere front to the cloud and rain of frontal regions. It is also desirable to have some idea regarding the means by which the frontal zone is created and maintained. It was with these objects in view that a series of flights was planned by the Meteorological Research Flight of the British Meteorological Office. Each flight was intended to cross the frontal zone at one or more levels in horizontal flight, and information on temperature, frost point and cloud were recorded.

In the present paper I wish to describe some of the observations, and discuss their interpretation in terms of the dynamics of the jet stream and frontal zone.

2. The observations

A Hastings aircraft was used for the flights and observations of temperature were made at half-minute intervals on a flat-plate electrical thermometer mounted beneath the nose of the aircraft. The error of the thermometer was probably about $\pm$ 0.2°F and some doubt (of the order of 1°F) attaches to the correction for airspeed which should have been applied when the aircraft was in cloud. However neither of these uncertainties can affect materially the temperature profiles shown in Fig. 1 and 2.

/Frost

Frost point was obtained by means of a manually operated Dobson-Brewer frost point hygrometer. The probable error of these observations is less than 3°F, but the records are not so complete as those of temperature because of the difficulty of maintaining a long series of readings at only ½-minute intervals.

Visual observations of cloud were recorded on the flights.

3. Temperature profiles

In order to examine the thermal structure of fronts in mid-troposphere the temperature observations made during all horizontal traverses of fronts at (or near) the 500 mb and 600 mb levels have been collected in Fig. 1 and 2. In these diagrams the temperature has been plotted against time but since the aircraft speed was usually around 220 kt and the flight track roughly perpendicular to the frontal zone, the plotted temperatures may be regarded as displaying the horizontal profile of temperature across the front with the scale of distance indicated at the foot of the diagram. One point of the temperature profile has been identified as corresponding to the warm boundary of the frontal zone, and the temperature plots have been aligned for convenience with these points in a vertical line. (All flights which intersected an identifiable front in mid-troposphere on a cross-section based on radiosonde observations have been included). Data for some of the flights have already been published (Sawyer, 1955); other observations come from flights not yet fully analysed. On some occasions level runs were made near both the 500 mb and 600 mb levels, but on other occasions only one level in mid-troposphere was included.

At 500 mb, a well-defined frontal zone with quite sharp changes of temperature gradient can be identified on about half of the occasions. On the remaining flights the region of temperature change is more or less diffuse. The width of the frontal zone as observed on these flights varied between 35 and 170 n.m., and the temperature change from about 6 to 18°F. The strongest maintained temperature gradient was that on 29th November, 1954 which amounted to 10°F in 35 n.m.

At 600 mb the temperature profiles are similar but, on the whole, rather less clearly defined. On less than a third of the temperature profiles in Fig.2 can the frontal zone be identified unequivocably. This may be purely an accident of the sample but, more probably, there is a greater tendency to form a narrow frontal zone with clear-cut boundaries at 500 mb.

4. Distribution of humidity

Fig. 3 and 4 show the horizontal variation of frost point depression at the 500 mb and 600 mb level. With one exception for which no hygrometer observations are available, these correspond to the temperature observations in Fig. 1 and 2. The frost point plots are aligned with respect to the warm boundary of the frontal zone as defined by the temperature observations.

The most striking feature of these humidity observations is the very dry air which is frequently, but not always, found in and near the frontal zone. On seven out of the nine available frost point traverses at 500 mb, the frost point depression exceeds 30°F and on two occasions frost point depressions exceeding 60°F were measured. On the flights which failed to encounter this very dry air, the air is still not really moist. Frost point depressions less than 10°F seem to be the exception and cloud in the frontal zone was encountered only once. On two other occasions the aircraft flew through cloud some distance on the warm side of the frontal zone, and once on the cold side. Cloud would no doubt have been encountered more frequently if the flights could have been extended for greater horizontal distances into the warm air.

/The

The distribution of the frost point depression does not appear to be closely related to the frontal zone or its boundaries, except that when the frontal zone was very clearly defined by the temperature profile (e.g., 29th November, 1954, 3rd November, 1950, 3rd November, 1953, 27th April, 1951) the very dry air was present and the frost point depression reached its extreme near the warm boundary of the frontal zone.

At 600 mb the humidity distribution seems even more erratic than at 500 mb. Some very dry air (frost point depression greater than 30°F) occurred on 7 out of the 12 flights and on most of them frost point depression is greater than 10°F over considerable stretches of the flight. On three of the flights the very dry air was closely associated with the frontal zone but on the remainder it was irregularly distributed.

The aircraft passed through cloud on 9 out of the 12 flights. On six of these the cloud did not significantly enter the frontal zone, being mainly on the warm side. On three flights, however, there was extensive cloud in the frontal zone and on one of these solid cloud appeared to fill most of the transitional region. This occurred in a broad warm-front zone with a steep slope of 1 in 30 and might be regarded as exceptional.

5. Summary of the observations

In order to summarise the data average temperature and frost point profiles have been calculated (Fig. 5 and 6). This has been done using the position of the warm boundary of the frontal zone as a reference point, and the temperature at this point as a reference temperature. In view of the wide dis-similarity of individual profiles the resulting average cannot be regarded as particularly representative. However, it serves to emphasise that at 500 mb the air in and surrounding the frontal zone is normally dry and has the lowest relative humidity near the warm boundary of the frontal zone. The average frost point depression everywhere exceeds 15°F and reaches 35°F in the frontal zone. There is no sign of a rapid increase in water-content across the frontal zone or its boundaries and it seems quite clear that at this level the frontal zone does not represent a transition from a cold dry air mass to a warm moist one; nor is the idea of up-gliding of air immediately above the frontal zone substantiated.

At 600 mb the average structure is somewhat different. The average frost point depression is relatively small just on the warm side of the frontal zone (less than 10°F), while a dry region occurs in the frontal zone (on which the frost point depression exceeds 20°F). The moist region just in the "warm air" adjacent to the front is found on individual flights and saturation was reached in this region on several occasions with accompanying cloud.

6. Some theoretical considerations

Fronts are closely linked with depressions and with the jet stream, and any completely satisfying explanation of the character of fronts would probably have to start with a theory of depressions and the jet stream. Nevertheless there is value in more limited theoretical discussions, and in the following paragraphs I have attempted to show that some of the characteristics of fronts in the upper troposphere follow as consequences if we accept the existence of the jet stream, and the fact that it has a coherent and only slowly varying structure. Thus I wish to side-step the question "Why is the front there?", and merely consider why some of the properties which we observe are necessary for its self-consistency.

For this purpose it is useful to consider the conditions under which a jet stream can persist in the upper troposphere in only slowly varying form.

/It

It is a fact of observation that this occurs, and that the jet stream current is sinuous and that the velocity in the core varies along its length. The troughs and ridges and the velocity maxima move considerably more slowly than the wind. For simplicity, we may regard the flow pattern as stationary with respect to the flow at some level in mid-troposphere. To be specific, we regard the flow as steady relative to co-ordinates moving with the 600 mb flow, and we examine the changes of vorticity which occur at higher levels as the air flows through the pattern.

Cross-sections drawn through the jet stream draw attention to the narrow character of the velocity maximum and the high values of horizontal wind shear, more particularly on the cold side. A typical wind profile would be that represented in Fig. 7(a) and the corresponding vorticity profile based on the wind shear would be as represented in Fig. 7(b).

Changes in the intensity of the jet stream or changes of its curvature along its length will involve changes in the vertical component of vorticity, and it is useful to examine how these arise. The procedure which I shall adopt is to assume (1) that the jet stream intensifies without change of wind profile or alternatively (2) that from being straight the jet becomes curved downstream (again without change of form), and I shall examine the pattern of divergence and vertical motion which would be necessary to satisfy the vorticity equation. Regarded strictly this procedure is invalid because the wind profile may, and probably does, change along the stream but the only alternative appears to be to attempt the solution of the three-dimensional dynamical problem. The simpler approach is both more practical, and shows more clearly the interaction of the various dynamical processes.

The vorticity equation in isobaric co-ordinates can be used to describe the change of vertical component of vorticity as the air flows along the jet stream. Thus

$$\frac{D\zeta}{Dt} = \frac{\partial \omega}{\partial y}\frac{\partial u}{\partial p} + \frac{\partial u}{\partial x}\frac{\partial v}{\partial p} + (f+\zeta)\,\mathrm{Div}_p \underset{\sim}{v} = 0 \qquad (1)$$

where $\omega \equiv Dp/Dt$

If we choose the x axis along the direction of the jet stream, $\partial u/\partial x$ is very much greater than $\partial v/\partial p$, and we can neglect the term $\frac{\partial \omega}{\partial x}\frac{\partial v}{\partial p}$

If, further, we assume that the flow is steady relative to some co-ordinate system moving with the general features of the flow at a speed corresponding to the wind at say 600 mb, and we ignore the local vorticity change in this moving co-ordinate system compared with the advective change, then

$$-V\frac{\partial \zeta}{\partial s} + \omega\frac{\partial \zeta}{\partial p} - \frac{\partial \omega}{\partial y}\cdot\frac{\partial u}{\partial p} + (f+\zeta)\,\mathrm{Div}_p \underset{\sim}{v} = 0 \qquad (2)$$

where V is the wind velocity relative to the wind at 600 mb and $\frac{\partial \zeta}{\partial s}$ is the vorticity change along a horizontal projection of a stream line.

In order to make some general deductions about the character of the flow we will assume that $V\frac{\partial \zeta}{\partial s}$ can be roughly estimated from the fact that the

jet is assumed (1) to intensify downstream or, alternatively, (2) to acquire curvature along its length. We may also assume from the presence of relatively undistorted isentropic surfaces in the stratosphere only a little above the jet that there is a level of small vertical motion somewhere near the jet level. At such a level the principal terms of equation (2) must be $V\partial\zeta/\partial s$ and $(f+\zeta)\mathrm{Div}_p \underset{\sim}{v}$. The divergence $\mathrm{Div}_p \underset{\sim}{v}$ can be calculated from these two terms alone.

On this basis, the distribution of $\mathrm{Div}_p \underset{\sim}{v}$ across the jet stream must be broadly as indicated in Fig. 7(c) and 7(d) for the two assumptions (1) of a jet stream intensifying along its axis and (2) of acquiring cyclonic curvature downstream. The particular conditions assumed in deriving Fig.7(c) and 7(d) respectively are (1) that the jet stream velocities increase by 10 per cent in 100 miles downstream $\left(V\frac{\partial\zeta}{\partial s} = \frac{V\zeta}{10^3}\ hr^{-2} \text{ with } V \text{ in kt}\right)$ and (2) that the jet stream acquires a curvature with 600 n.m. radius at a distance 1000 n.m. downstream from a straight section.

An interesting feature of these distributions of divergence is the strong gradients of divergence across the jet stream, and in particular the strong increase of convergence towards the jet axis from the cold side. If the vertical velocity is small near the jet stream level, then at lower levels towards mid-troposphere a strong gradient of vertical velocity must be expected with strong descent near to the jet stream axis and relatively weak descent only some 200 miles further towards the cold side of the jet. This differential subsidence can be expected to be effective in setting up a strong horizontal gradient of temperature in mid-troposphere as observed in the 500 mb frontal zone.

However the picture is more complicated, because the resulting values of the vertical velocity are large enough to make the second and third terms in equation (2) comparable in magnitude with the others. (The second and third terms represent respectively the vertical advection of vorticity and the vorticity change arising from the tilting of the vortex tubes). Were in fact the values of the divergence represented in Fig. 7(c) and 7(d) continued down to the 500 mb level then there would be values of ω and $\partial\omega/\partial x$ occurring at that level of as much as 10 mb/hr and 3.5×10^{-2} mb/hr^{-1}nm^{-1}. If we assume that the wind increases linearly from a uniform value of 40 kt at the 600 mb level up to the level of the jet stream at say 300 mb, then the order of magnitude of the terms in equation (2) would be as indicated in Table I. The crude assumption has been made that values of ω can be calculated from the divergence obtained ignoring terms $\omega\partial\zeta/\partial p$ and $-\partial\omega/\partial y \cdot \partial u/\partial p$.

Table I - Relative magnitude of terms in equation (2)

		$V\frac{\partial\zeta}{\partial s}$	$+\omega\frac{\partial\zeta}{\partial p}$	$-\frac{\partial\omega}{\partial y}\frac{\partial u}{\partial p}$	$+(f+\zeta)\mathrm{Div}_p V$
		(Units hr^{-2})			
400 mb	50 mi. from axis	1.6×10^{-2}	-0.4×10^{-2}	0	-1.2×10^{-2}
	150 mi. from axis	0.7×10^{-2}	-0.2×10^{-2}	-0.2×10^{-2}	-0.3×10^{-2}
500 mb	50 mi. from axis	0.4×10^{-2}	-0.8×10^{-2}	0	$+0.4 \times 10^{-2}$
	150 mi. from axis	0.2×10^{-2}	-0.4×10^{-2}	-0.4×10^{-2}	$+0.6 \times 10^{-2}$

/It

It is immediately clear that both the vertical advection of vorticity, $w\,\partial\zeta/\partial p$, and the "tilting term" $-\partial w/\partial y \cdot \partial u/\partial p$, are likely to be significant in the region immediately below and on the cold side of the jet stream. This has been confirmed, particularly as regards the latter term, by individual case studies by Reed and Sanders (1953), Reed (1955) and Newton (1954).

The values of ω on which the preceding figures are based assume that the convergence can be calculated ignoring the two terms which we have seen to be important. This is clearly invalid. In fact we can deduce that the convergence rapidly decreases downwards, and may even be replaced by divergence. It is also clear that this reversal should take place more rapidly at 150 miles from the jet axis than closer to the axis because the stronger horizontal gradient of ω at the former position gives rise to larger values of the "tilting term". The fields of divergence and of vertical motion in the region below and on the cold side of the jet must be roughly as indicated in Fig. 8(a) and (b).

The region of horizontal divergence below and to the cold side of the jet stream will bring about an intensification of the vertical gradient of potential temperature (i.e., it will increase the vertical stability) in this region as is observed in the 500 mb frontal zone. At the same time the tilting about a horizontal axis will increase the horizontal temperature gradient, and this will be accompanied by an increase in the cyclonic vorticity about a vertical axis apparently primarily as a result of the "tilting term", $-\partial w/\partial y \cdot \partial u/\partial p$, in equation (1).

It is not profitable to pursue the argument further without a more satisfactory treatment of the balance between the various dynamical factors involved. Nevertheless, the treatment has probably been adequate to show that a jet stream which has a strong shear on its cold side, and flows from a region of relatively low speed to one of higher speed, must develop a region of general subsidence on its cold side which is accompanied by an intensification of both horizontal and vertical gradients of potential temperature such as is observed in 500 mb frontal zones.

Similar conclusions would follow regarding regions where the jet stream flows into a slow-moving upper trough although the frontal zone would then be built up at a smaller horizontal distance from the jet axis.

This dynamical process provides no explanation as to why the boundaries of frontal zones should be so sharp and well-defined as is often observed. However, these rather abrupt changes of temperature gradient may well arise from the intensification and tilting of already existing boundaries to inversions or stable layers. The horizontal divergence will intensify the changes in vertical gradients of temperature and the tilting will introduce into the horizontal temperature gradients much stronger discontinuities than are normally present. Reed and Sanders (1953) describe a case in which the subsidence was sufficiently great to bring stratospheric air into the frontal zone, and the tropopause itself provided the initial discontinuity of lapse rate which became the boundaries of the frontal zone. I also examined a similar case which occurred over north France on 3rd November, 1950 and which appears in Fig. 1 and 3 (Sawyer, 1955). However, these cases are probably exceptional, and stable layers and inversions in the troposphere may also provide the initial discontinuities of lapse rate which are intensified into the boundaries of frontal zones.

So far, I have discussed the formation of a frontal zone in intense sections of the jet stream and in the regions of upper troughs. The arguments may be reversed to suggest a reverse circulation and frontolysis as the air passes into weaker sections of the jet and into upper ridges. No doubt detailed observations of the jet stream and studies of upper fronts have concentrated on regions where they are intense, but I think

/there

there is another reason why the structure which is built up in these intense regions of the jet and frontal zone should persist elsewhere. That is because it is unlikely that vertical circulation as the air enters a diffluence or a ridge will be the exact reverse of that in a preceding confluence or trough. It will not be the same air which descended in the confluence that ascends again in the diffluence - sufficient small changes in the relative position of front and jet will probably have taken place to mean that the air which rises most will not be precisely that which descended most. Thus in association with weaker sections of the jet and with upper ridges, we might expect to find some remnant, albeit less intense, of the dry frontal zone built up in preceding a confluence or trough.

Conclusions

The temperature observations on horizontal traverses beneath the jet stream confirm that a narrow "frontal zone" some 30 to 50 miles wide and with sharp boundaries is often present at the 500 mb level. On other occasions the temperature changes are more diffuse. Humidity observations indicate that the air in and near the "frontal zone" has undergone subsidence and make it seem unlikely that the "frontal zone" at this level owes its origin to the different geographical origin of the air on either side of it. Theoretical considerations suggest the horizontal temperature gradient arises from the differential subsidence of air on the cold side of the jet stream which leads to tilting of the isentropic surfaces. Dynamical considerations also explain the vertical stability and cyclonic vorticity associated with the front.

The front-forming processes in the upper troposphere only operate where the jet stream is intensifying or becoming more cyclonically curved along its length, but the characteristic structure is carried forward into other weaker sections of the jet and probably does not degenerate completely. This may account for the variety of temperature and humidity profiles observed.

At 600 mb the temperature gradients at fronts are somewhat less well-defined than at 500 mb, and although dry subsided air is often found in the frontal zone, the air on the warm side of the warm frontal boundary is usually relatively moist. Saturation and cloud occur in this region and the structure corresponds more closely at 600 mb to the conventional front with upsliding warm air. This conforms with previous findings (Sawyer and Dinsdale, 1955) that the boundary of the frontal cloud mass slopes more steeply than the front itself and on the average intersects the warm boundary of the frontal zone around 600 mb.

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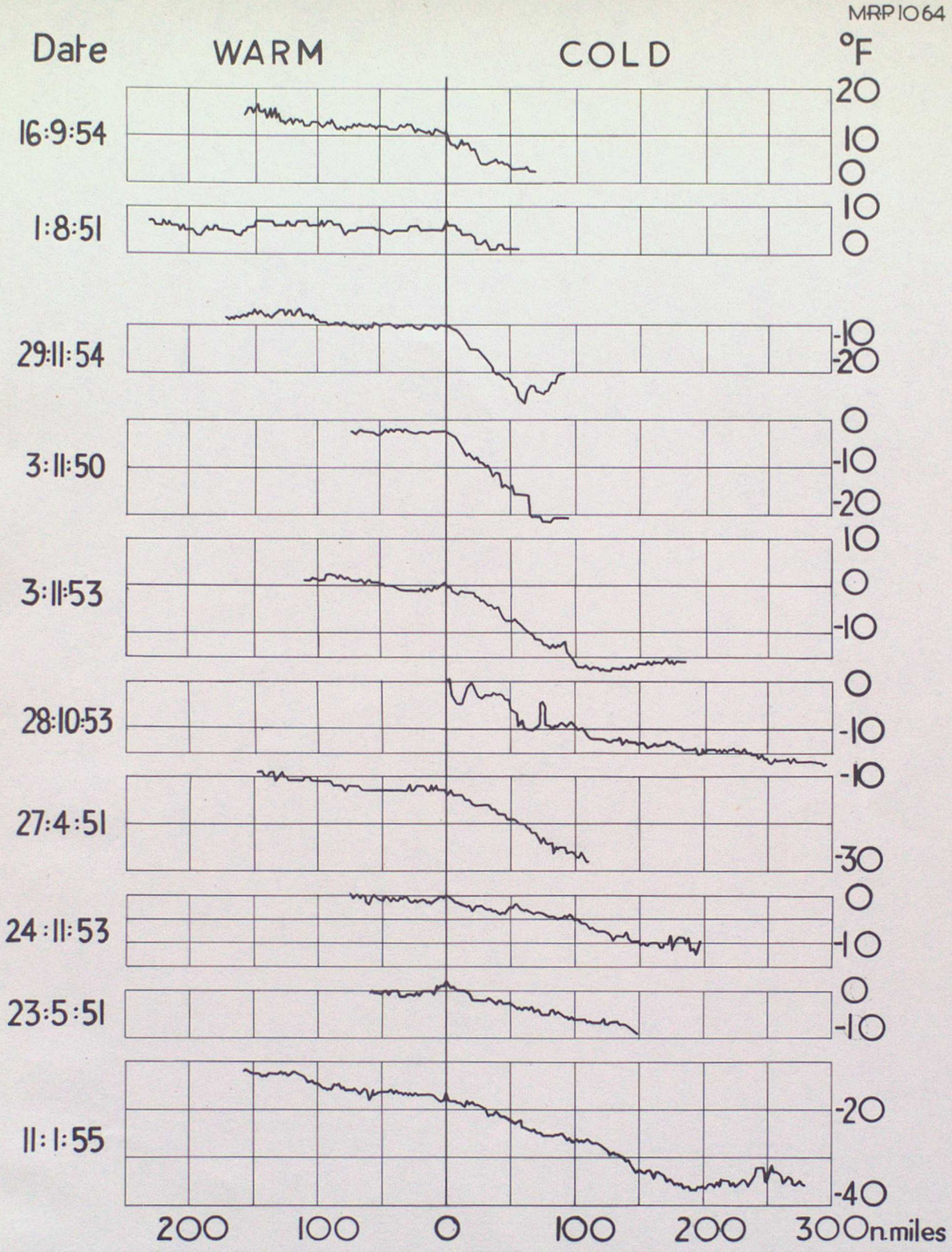
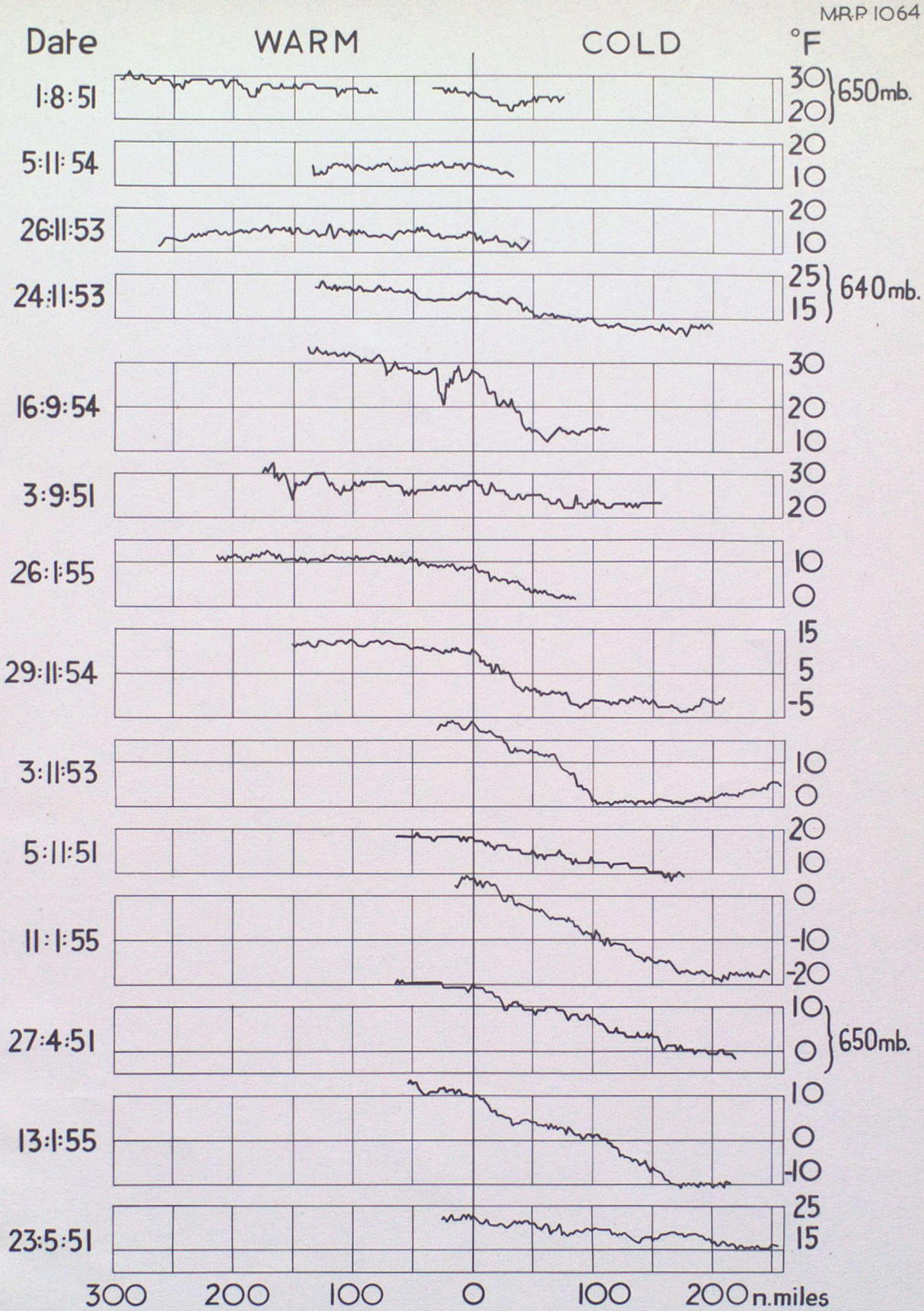


Horizontal distribution of temperature through frontal zones at 500mb.

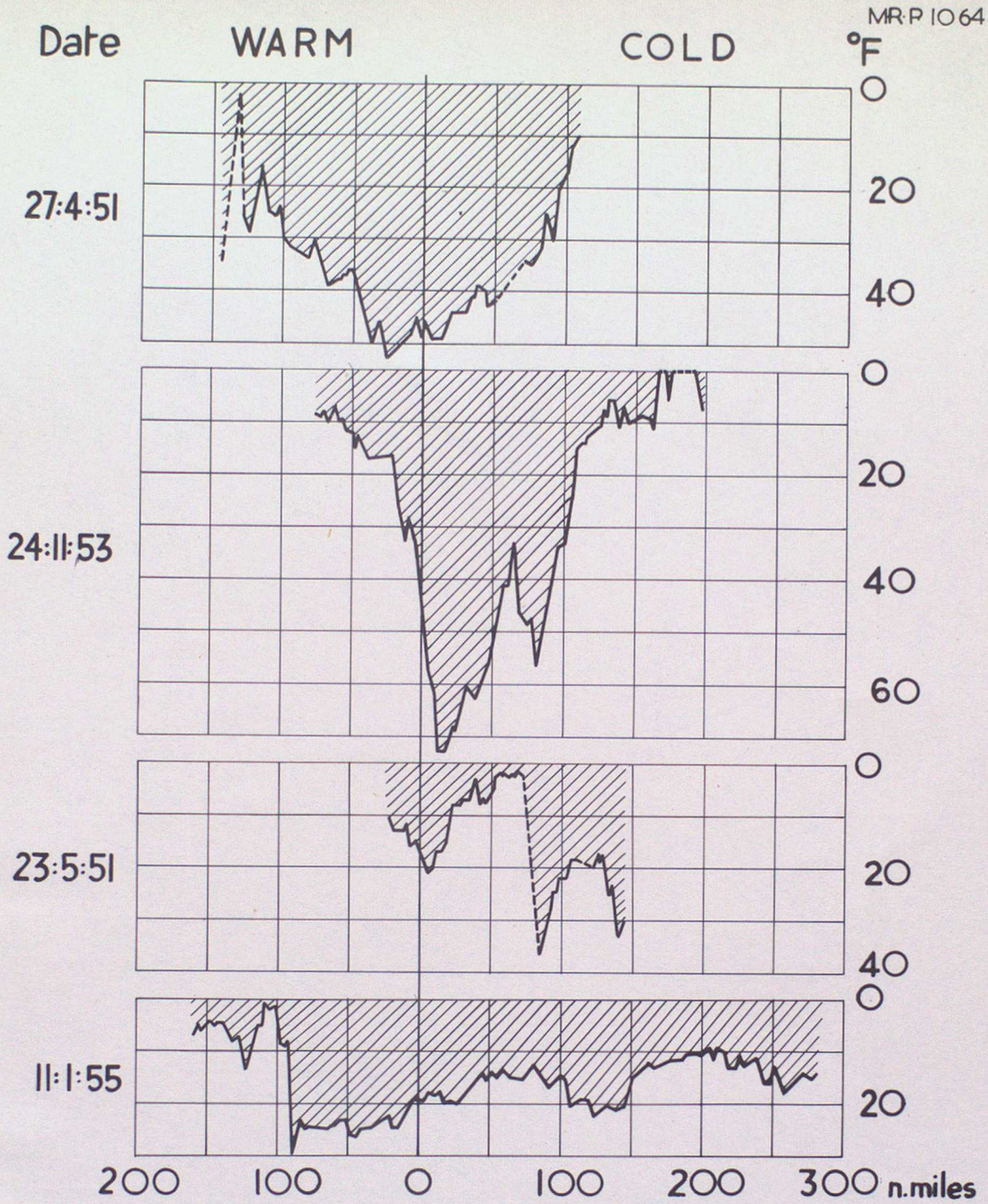
Fig. 1



Approximate scale of distance

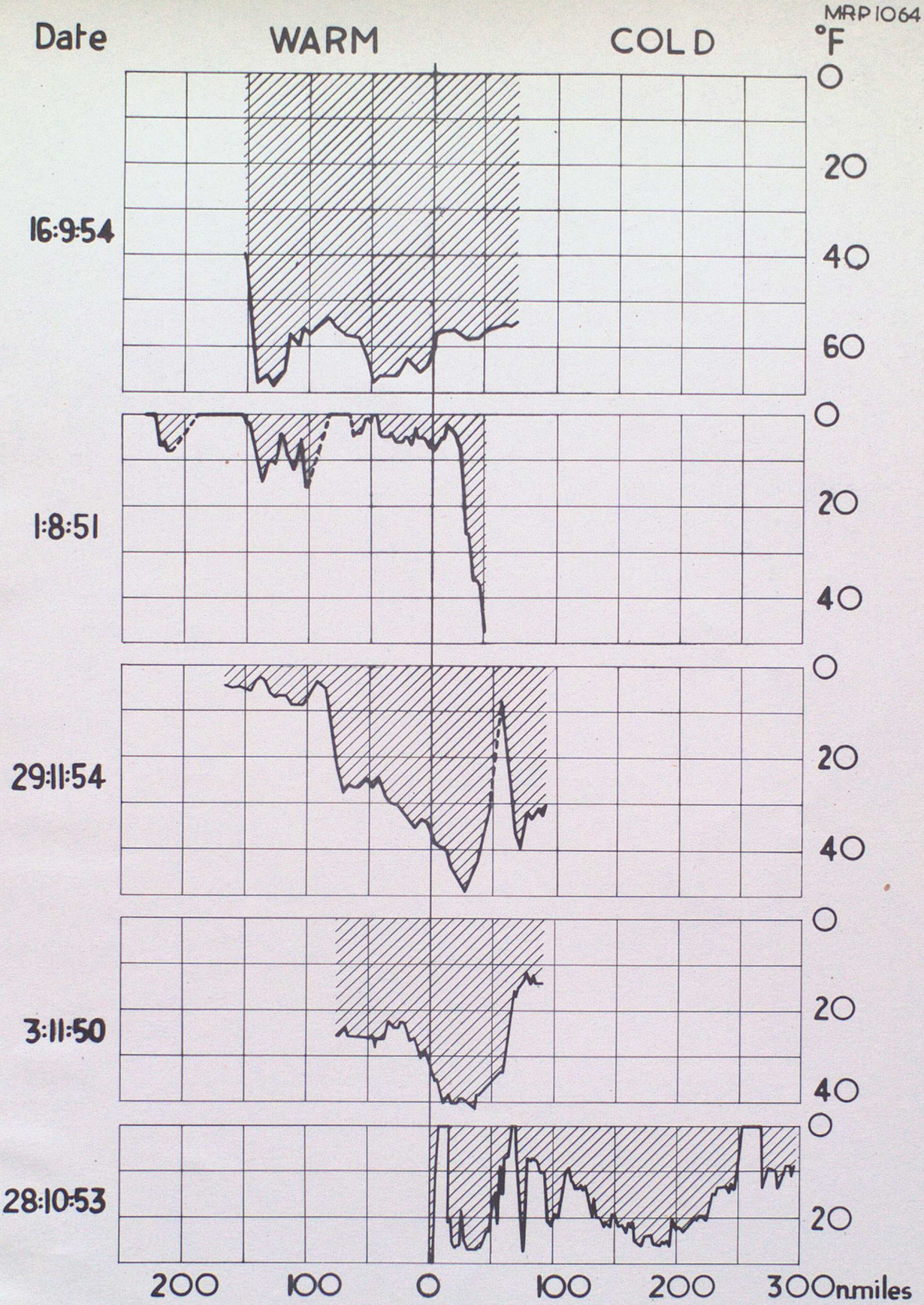
Horizontal distribution of temperature through frontal zones at 600mb.

Fig. 2



Horizontal distribution of frost point depression through frontal zones at 500 mb. (____ cloud)

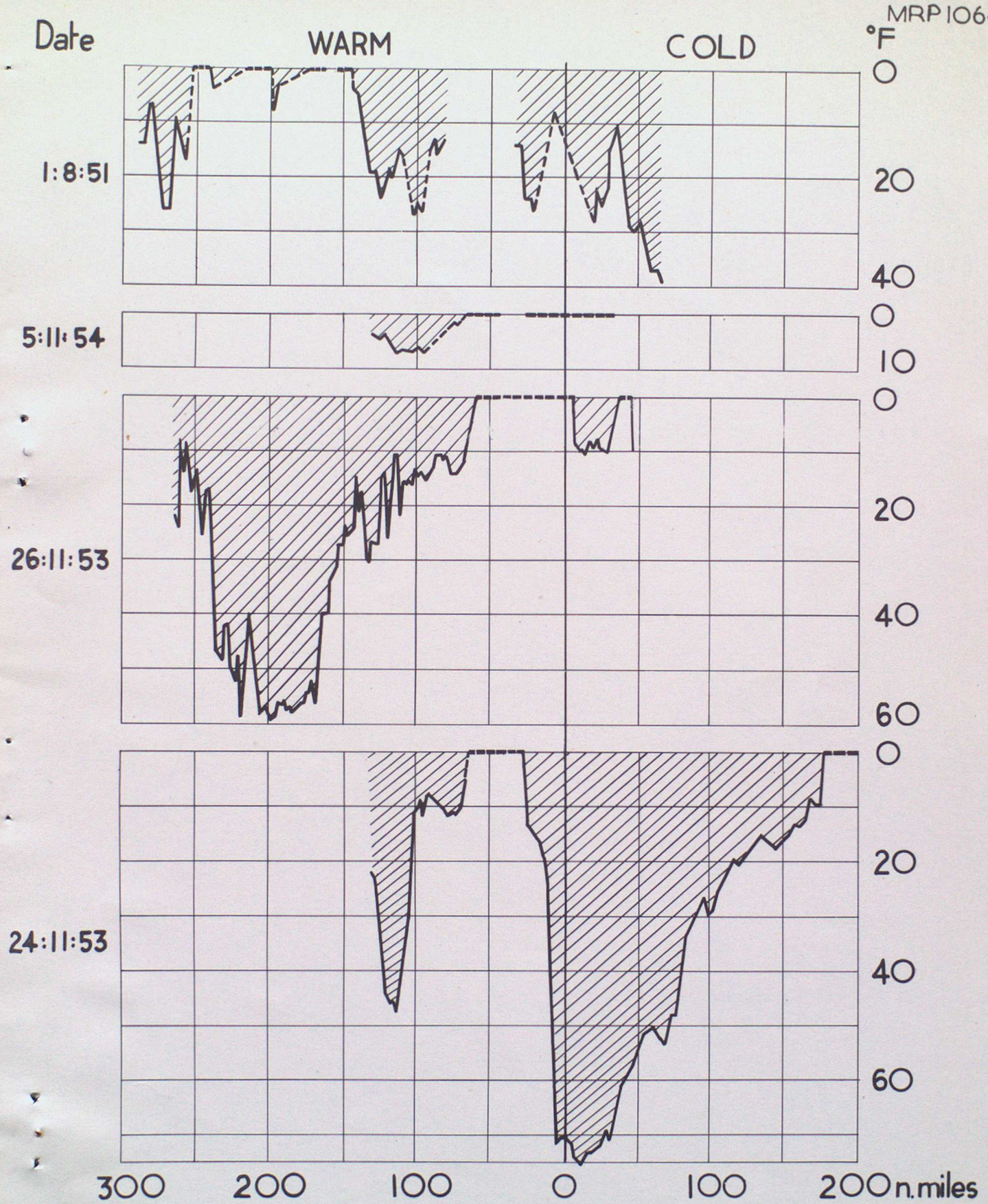
Fig. 3(b)



Approximate scale of distance

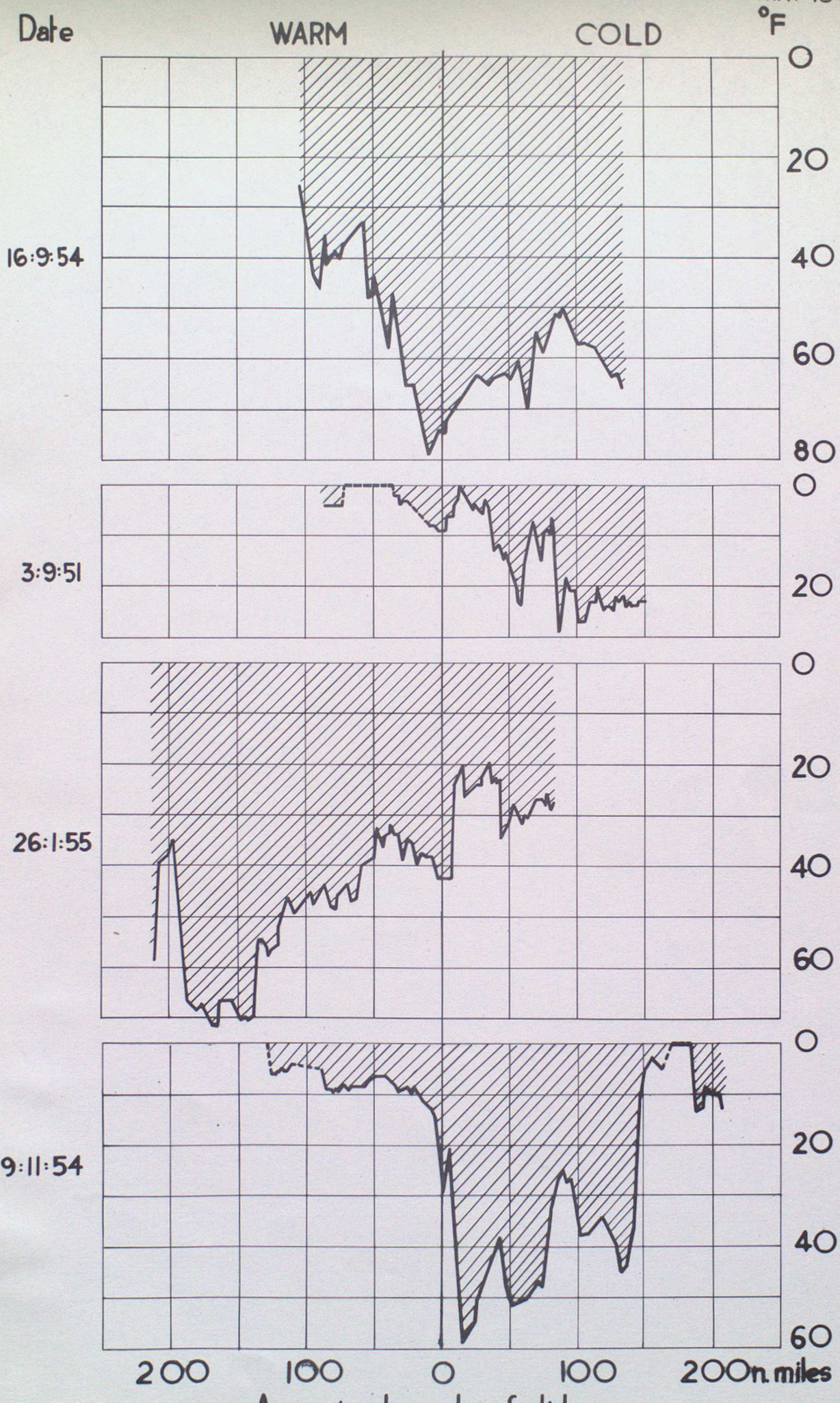
Horizontal distribution of frost point depression through frontal zones at 500 mb. (___ cloud)

Fig. 3(a)



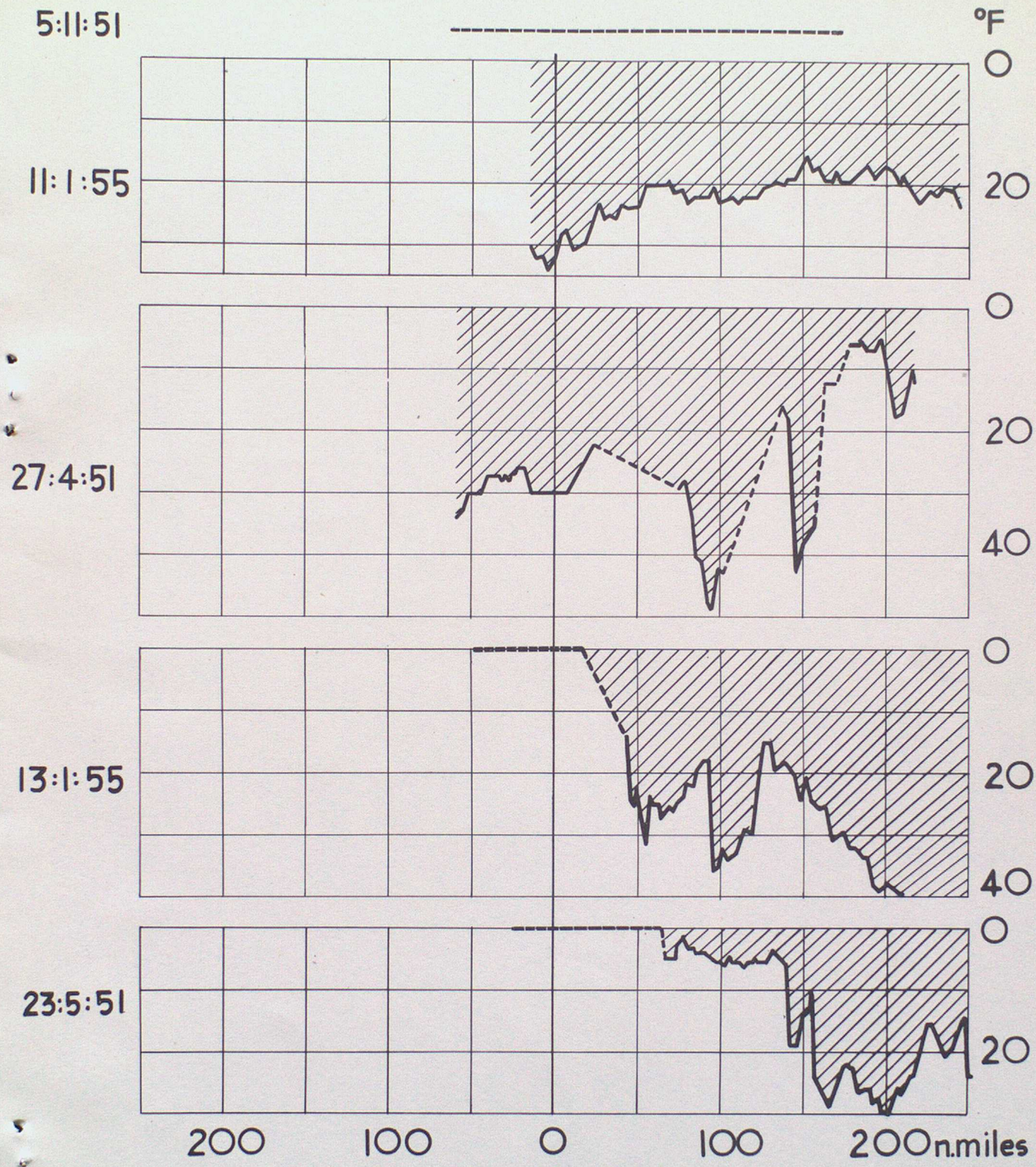
Horizontal distribution of frost point depression through frontal zones at 600 mb. (____ cloud)

Fig. 4 (a)



Horizontal distribution of frost point depression through frontal zones at 600 mb. (___ cloud)

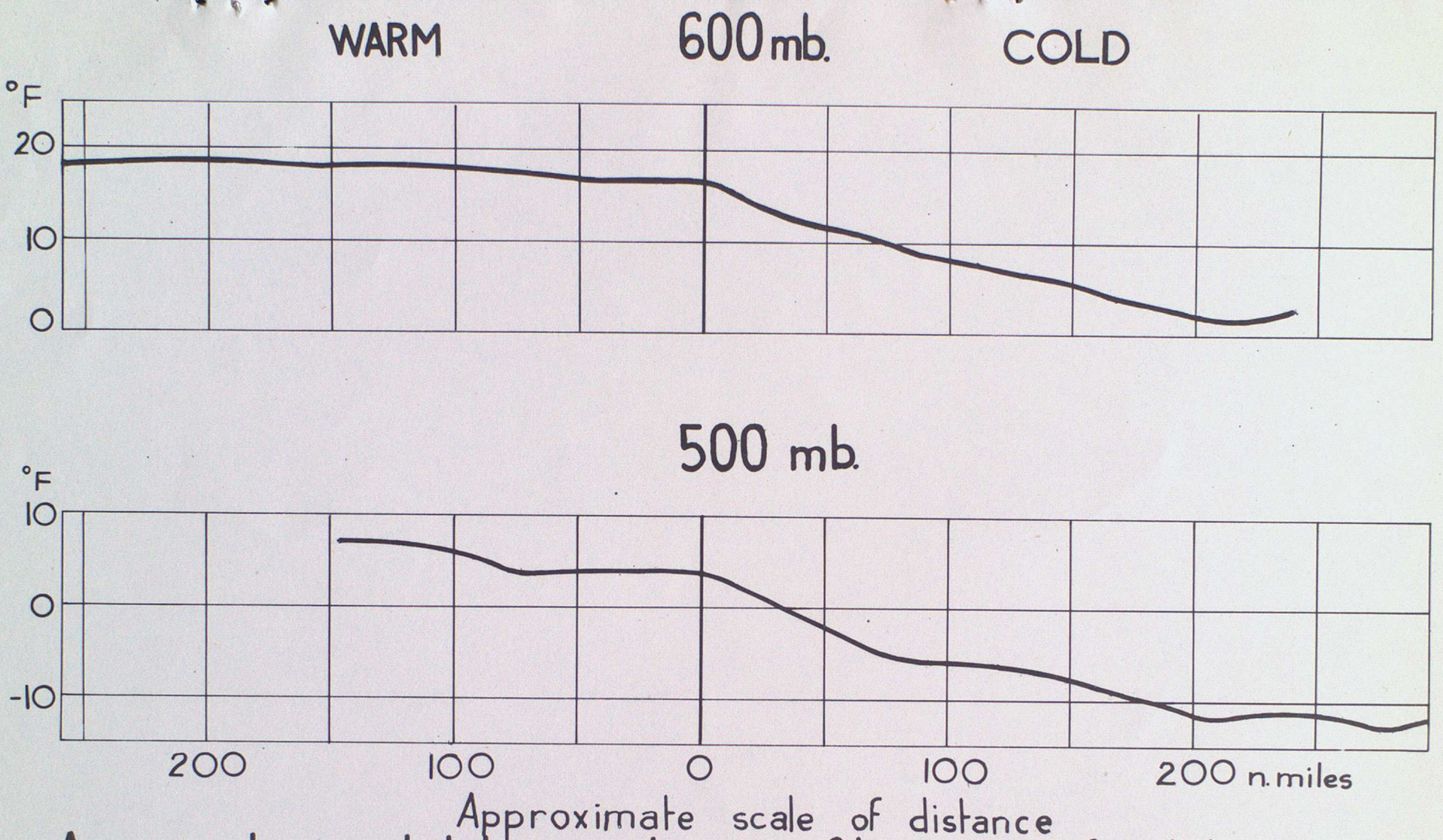
Fig. 4(b)



Approximate scale of distance

Horizontal distribution of frost point depression through frontal zones at 600 mb. (____ cloud)

Fig. 4 (c)



Average horizontal temperature profile across frontal zones at 500 mb. and 600 mb.

Fig. 5

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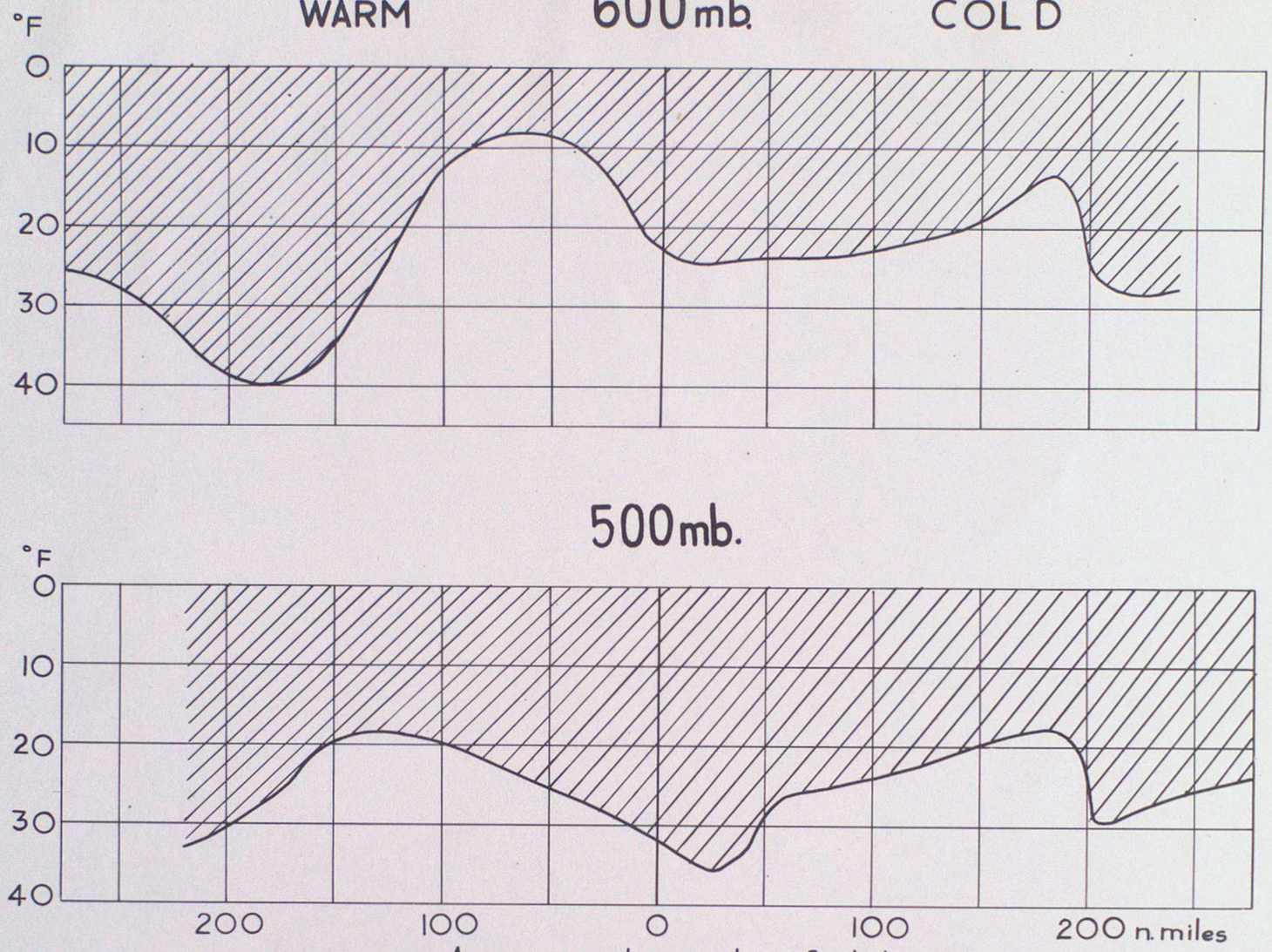
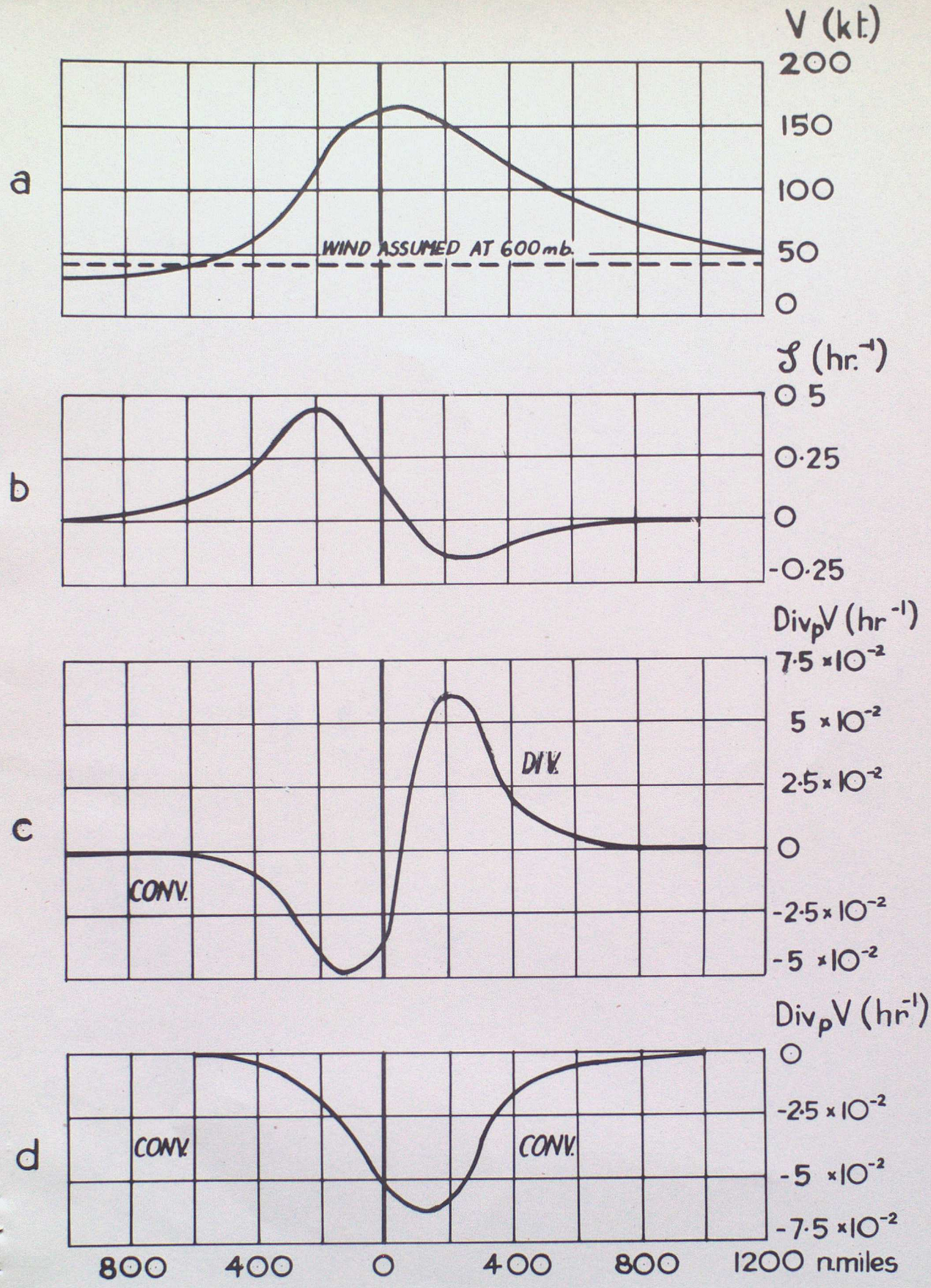


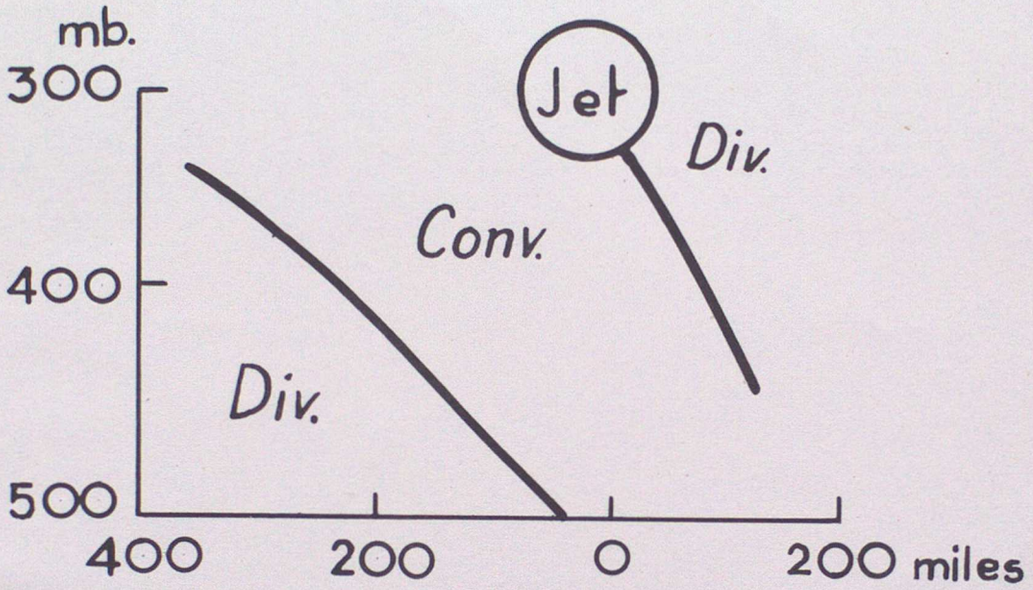
Average horizontal profile of frost point depression across frontal zones at 500mb. and 600mb.

Fig. 6

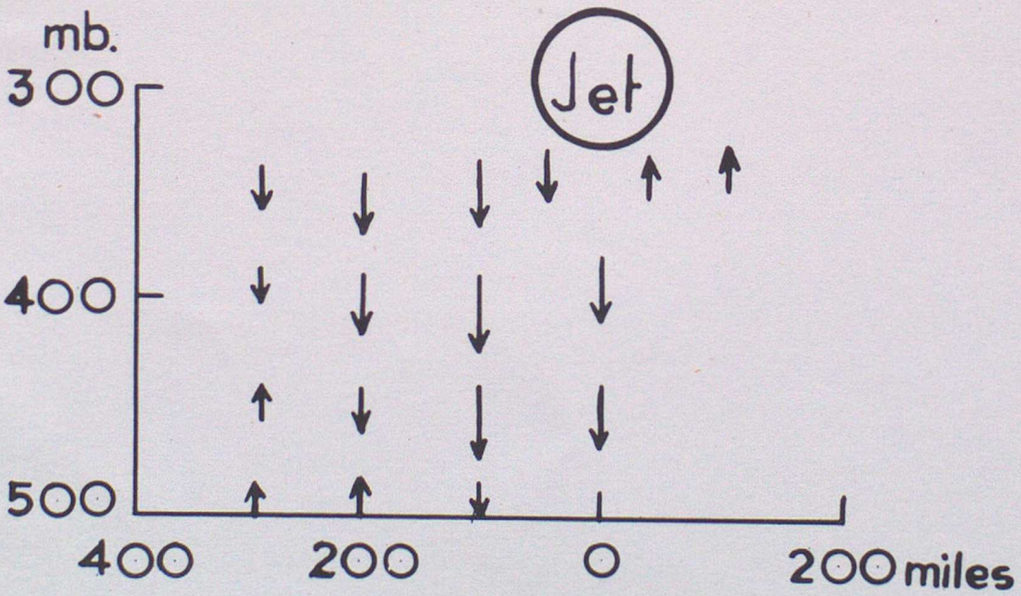


(a) Schematic profile of wind through a jet stream
(b) Corresponding profile of relative vorticity
(c) Distribution of divergence required if the jet stream intensifies along its length
(d) Distribution of divergence required if the jet stream is to acquire a curvation of 600 n. miles

Fig. 7



(a) Horizontal divergence



(b) Vertical motion

Distribution of horizontal divergence and vertical motion on the cold side of an intensifying jet stream

Fig. 8